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1 Inferring Interseismic Coupling along the Lesser 2 Antilles Arc: a Bayesian Approach

3 E. van Rijsingen¹, E. Calais^{1,2}, R. Jolivet^{1,2}, J. B. de Chabaliér³, J. Jara¹, S.
4 Symithe⁴, R. Robertson⁵, G.A. Ryan^{5,6}

5 ¹École Normale Supérieure, Department of Geosciences, PSL Université, CNRS UMR 8538, Paris, France

6 ²Institut Universitaire de France, 1 rue Descartes, 75005 Paris, France

7 ³Institut de Physique du Globe de Paris, Université de Paris, CNRS UMR 7154, Paris, France

8 ⁴URGéo Laboratory, State University of Haiti, Port-au-Prince, Haiti

9 ⁵Seismic Research Center, University of the West Indies, St. Augustine, Trinidad and Tobago

10 ⁶Montserrat Volcano Observatory, Flemmings, Montserrat, West Indies

11 **Key Points:**

- 12 • We use geodetic observations to estimate interseismic coupling of the Lesser An-
13 tilles subduction zone using a Bayesian approach
- 14 • We find low to very low interseismic coupling, making it less likely that the 1839
15 and 1843 historical earthquakes were thrust events
- 16 • The GPS data also shows small, but detectable along-arc extension, consistent with
17 observations of active normal faulting within the arc

Corresponding author: Elenora van Rijsingen, e.m.vanrijsingen@gmail.com

Abstract

The Lesser Antilles subduction zone is a challenging region when it comes to unraveling its seismogenic behavior. Over the last century, the subduction megathrust has been seismically quiet, with no large thrust event recorded, which raises the question whether this subduction zone is able to produce large interplate earthquakes or not. However, two historical earthquakes in the 19th century, a M 7-8 in 1839 and M 7.5-8.5 in 1843, are proposed to have occurred along the subduction megathrust, although no direct evidence exists. Here we provide a new assessment of interseismic coupling for the Lesser Antilles subduction zone, based on updated GPS velocities and the latest models of the slab geometry and elastic crustal structure. We use a Bayesian approach, allowing us to explore the entire range of plausible models and to provide realistic estimates of interseismic coupling and associated uncertainties. We find low to very low coupling along the entire plate interface, including in the proposed rupture areas of the 1839 and 1843 events, where the sensitivity of our model is high. While a further understanding of temporal variations in interseismic coupling needs to be addressed in future studies, our results indicate that the Lesser Antilles subduction zone is uncoupled, which challenges the idea that the 1839 and 1843 earthquakes were thrust events. The updated GPS velocities of this work now also reveal a small, but detectable amount of along-arc extension, consistent with geological observations of active normal faulting within the arc.

Plain Language Summary

The Lesser Antilles subduction zone forms the boundary between the North- and South American plates that sink underneath the overlying Caribbean plate. Such downgoing movement typically results in the buildup of stress along the frictional interface between the plates. When these stresses overcome the strength of the plate interface, they can be released through earthquakes, that may have devastating effects on societies. By using measurements from GPS stations on the islands of the Lesser Antilles, we aim to determine how much strain is currently being accumulated along the subduction interface, that is, how coupled the interface is. A high degree of coupling means that large “megathrust” earthquakes are likely, while a low coupling means that they are less likely and/or very rare. Two large earthquakes struck the Lesser Antilles in the 19th century, which have been interpreted to have occurred on the subduction interface. In this work we find a low to very low coupling along the plate interface, which implies that (1) these historical earthquakes are unlikely to have occurred on the subduction interface but rather deeper within the downgoing plate, and (2) large, Tohoku-like, megathrust earthquakes are unlikely and/or must be very rare in the Lesser Antilles.

1 Introduction

An important, but originally unexpected, outcome of geodetic measurements at subduction plate boundaries over the past 20 years is that some are locked, therefore building-up elastic strain to be released in large ($M_W > 7.5$) megathrust earthquakes, while others appear to slip aseismically at a rate close or equal to the plate convergence rate, without generating large events. The northern Honshu subduction zone in Japan is an example of the former, with a mechanically locked plate interface and elastic strain accumulation measurable on land, as documented in the decades preceding the March 11, 2011, $M_W 9.0$ Tohoku-Oki earthquake (Loveless & Meade, 2010, 2011; Mazzotti et al., 2000). The South Ecuador – North Peru segment of the South American subduction zone is an example of the latter, with a lack of large historical earthquakes and of elastic strain accumulation, indicative of a plate interface that is mechanically uncoupled (Nocquet et al., 2014). The development of geodetic networks has provided crucial information that allows us to map with some detail the spatial – and sometimes temporal – variability in interplate coupling at subduction zones (Chlieh et al., 2008, 2011; Freymueller & Beavan, 1999; Freymueller et al., 2000; Metois et al., 2016; Villegas-Lanza, Chlieh, et al., 2016; Villegas-Lanza, Nocquet, et al., 2016). Imaging, and understanding, the relationship between the degree of coupling of subduction plate boundary segments and their ability to produce – or not – megathrust earthquakes is of utmost importance to inform regional seismic hazard assessment (e.g., Loveless & Meade, 2011; Stevens & Avouac, 2016).

Subduction parameters proposed to play a role in tuning the seismogenic behaviour of the megathrust include convergence velocity and slab age (Peterson & Seno, 1984; Ruff & Kanamori, 1980), seismogenic zone width and trench-parallel extent (Brizzi et al., 2018; Schellart & Rawlinson, 2013), upper plate strain (Heuret et al., 2011, 2012), trench curvature (Schellart & Rawlinson, 2013), internal density contrasts (Song & Simons, 2003), curvature of the downgoing plate (Bletery et al., 2016; Schellart & Rawlinson, 2013), trench sediment thickness (Heuret et al., 2012; Scholl et al., 2015), and subduction interface roughness (Das & Watts, 2009; van Rijsingen et al., 2018). Although some of these parameters partially correlate with the global distribution of subduction megathrust earthquakes, some subduction zones remain poorly understood, in particular those that have been seismically quiet over the instrumental time period. Such regions are not devoid from significant events, but are referred to as quiet because no large thrust event has been recorded in the instrumental, and sometimes historical, period. To better understand the long-term seismogenic behaviour of such quiet subduction zones, one must therefore rely on geological and historical records of earthquakes, as well as interseismic coupling estimates

88 inferred from geodetic measurements (Hough, 2013; Satake & Atwater, 2007; Wang &
89 Tréhu, 2016).

90 The Lesser Antilles subduction zone is one of these quiet subduction zones, with
91 no thrust events larger than M_w 6.5 observed within the instrumental time interval. In
92 fact, the four largest earthquakes recorded in the past 100 years (M_S 7.5 1953; M_S 7.5
93 1969; M_S 7.4 1974; and M_W 7.4 2007) were all the result of normal faulting within the
94 subducting slab or overriding plate (e.g., McCann et al., 1982). However, two large his-
95 torical events in the 19th century, a M 7-8 event in 1839 and a M 7.5-8.5 in 1843 have
96 been interpreted by some as interplate thrust events, although no direct evidence exists
97 (Bernard & Lambert, 1988; Feuillet et al., 2011; Hayes et al., 2014; Hough, 2013). If con-
98 firmed, this would be an indication that similar large interplate thrust earthquakes are
99 to be expected in the future. In that case, according to current models of subduction zone
100 seismogenic behavior showing that fault locking is a stable feature over at least 1000s
101 of years (Song & Simons, 2003; Avouac, 2015; Mouslopoulou et al., 2016; Jolivet et al.,
102 2020), it is reasonable to assume that 175 years after such large thrust events the Lesser
103 Antilles plate interface should have relocked and that elastic strain accumulation should
104 be visible in present-day surface deformation measurements.

105 Early GPS measurements in the Caribbean showed that a geodetic site on Barba-
106 dos island, well within the area that should experience elastic strain accumulation if the
107 plate interface was locked, was moving at a velocity consistent with that of the Caribbean
108 plate (DeMets et al., 2000), indicative of very low coupling on the interface. Since then,
109 thanks to the rapid development of geodetic observations in the Lesser Antilles, two stud-
110 ies have attempted to estimate interseismic coupling along the subduction interface (Manaker
111 et al., 2008; Symithe et al., 2015), both finding very low values. However, uncertainties
112 related to the distance of the GPS-stations from the trench, the non-uniqueness of the
113 inversion, the crude estimation of coupling uncertainties and a limited data set all war-
114 rant a revision of this work with better data and a more advanced inversion technique.

115 In this study, we therefore determine the degree of interplate coupling on the Lesser
116 Antilles subduction using updated GPS velocities and more accurate models of the slab
117 geometry and the elastic structure of the crust, while adopting a Bayesian inversion ap-
118 proach. By exploring the entire range of model parameters, this approach provides an
119 estimate of the interseismic coupling together with a probabilistic measure of its uncer-
120 tainty. Our goal is to shed more light on the seismogenic behaviour of the Lesser An-
121 tilles subduction and to discuss what this could mean for seismically quiet subduction
122 zones in general. How does their short-term behavior relate to their ability to rupture

123 in large megathrust earthquakes, and is there a physical mechanism that can explain their
 124 long-term aseismic character?

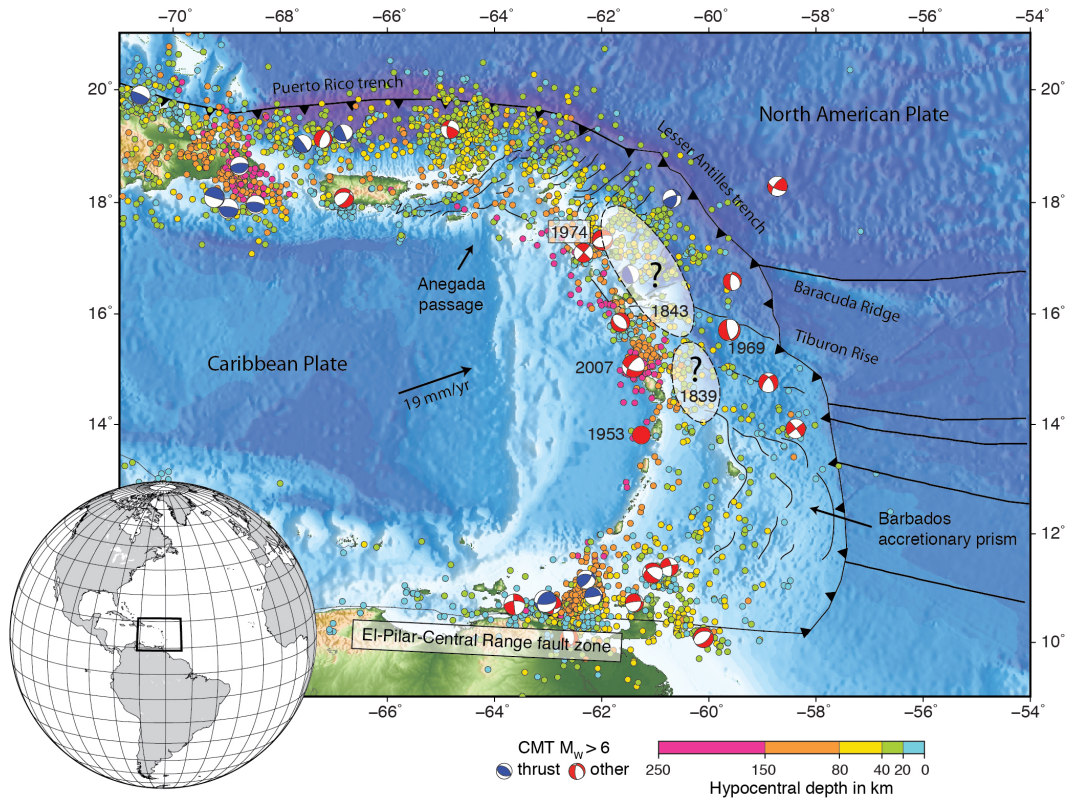


Figure 1. Seismotectonic setting of the Lesser Antilles subduction zone. Colored circles indicate seismicity (M_W 4-6) from the USGS catalog, color coded as a function of depth. Global CMT Catalog (1976-2020) focal mechanisms are plotted in red and blue ($M_W > 6$). The white shaded areas represent the proposed rupture areas of the 1839 and 1843 historical earthquakes (e.g., Feuille et al., 2011). The thin black lines indicate the faults mapped by Feuille et al. (2002).

125 2 Tectonic Setting

126 The intra-oceanic Lesser Antilles subduction zone forms the eastern boundary of
 127 the Caribbean plate (Figure 1). Since the Eocene, Atlantic oceanic crust of both the North-
 128 and South American plates has been subducting westward at a slow convergence rate
 129 of 18-20 mm/year (DeMets et al., 2010). The Lesser Antilles arc is bounded to the north
 130 by the Anegada passage, an extensional fault system, also marking the eastern end of
 131 the Greater Antilles (Jany et al., 1990; Laurencin et al., 2017; Masson & Scanlon, 1991).

132 To the south, the Lesser Antilles arc abuts against the right-lateral El-Pilar-Central Range
133 strike-slip fault zone that marks the boundary between the Caribbean- and South Amer-
134 ican plates (Mann et al., 1990). With an azimuth of $\sim 251^\circ$, the subduction direction is
135 almost arc-perpendicular in the center of the arc, while becoming more oblique towards
136 the northern and southern edges.

137 As the subduction becomes more oblique in the north, the arcuate slab changes from
138 dipping to the west underneath the Lesser Antilles, to plunging to the south below His-
139 paniola and Puerto Rico (Masson & Scanlon, 1991; McCann & Sykes, 1984). The tran-
140 sition between the North American and South American plates has been proposed to oc-
141 cur around 15° , where the existence of a slab gap at depth is debated (van Benthem et
142 al., 2013; Patriat et al., 2011; Pichot, 2012; Schlaphorst et al., 2017). According to two
143 recent models of the Lesser Antilles slab geometry, the shallow slab dip changes from $\sim 14^\circ$
144 in the north, to a shallower angle of $\sim 7^\circ$ towards the south (Bie et al., 2020; Hayes et
145 al., 2018). Below the arc, the slab dips much more steeply, with some differences between
146 the different slab models. For instance, in the central part of the subduction zone, the
147 global Slab2 model (Hayes et al., 2018) estimates the slab surface to be up to 70 km shal-
148 lower than the Bie et al. (2020) model.

149 The 850-km-long Lesser Antilles volcanic arc consists of 11 major volcanic islands
150 and 19 small islands (the Grenadines) between St. Vincent and Grenada in the south.
151 The arc is constructed on thickened, > 150 Ma old oceanic crust of the Caribbean plate
152 (Mauffret & Leroy, 1997), with estimates of crustal thickness varying between 21 and
153 35 km (Bie et al., 2020; Gonzáles et al., 2018; Schlaphorst et al., 2018). North of Mar-
154 tinique, the arc splits into two branches, with the inner arc (containing St Kitts, St Eu-
155 statius and Saba) still volcanically active today. The islands of Antigua and St Martin
156 are part of the remnants of the inactive outer arc (Bouysse & Westercamp, 1990). To-
157 wards the south, the arc becomes narrower, more continuous and contains fewer volcanic
158 islands (Feuillet et al., 2002).

159 With its slow convergence rate and old subducting lithosphere (80-100 Myr), the
160 Lesser Antilles subduction zone is a global end-member (Stein et al., 1983). It is also an
161 end-member in terms of incoming plate structure, as it consumes slow-spreading (2 cm/yr)
162 Atlantic lithosphere, while Pacific subduction zones consume much faster-spreading (up
163 to 15 cm/yr) oceanic lithosphere (Müller et al., 2008). Several fracture zones, well-marked
164 in the bathymetry, are entering the trench, as well as two elongated bathymetric highs,
165 the Baracuda Ridge and the Tiburon Rise (Bouysse & Westercamp, 1990; McCann &
166 Sykes, 1984; Stein et al., 1982), now interpreted as compressional structures within the
167 ~ 200 km wide transition zone between the North- and South American plates (Patriat

168 et al., 2011; Pichot, 2012). The sedimentary cover entering the subduction shows large
169 variations in thickness and nature along the arc. In the south, the large influx from the
170 Orinoco river built a 7-km-thick layer of mainly continental clastic sediments, that con-
171 tributes in building the Barbados accretionary prism (Speed & Larue, 1982). North of
172 the Barracuda ridge, the seafloor is covered by only 200 m of dominantly pelagic ma-
173 rine sediments (Reid et al., 1996).

174 The forearc structure also shows a transition from north to south (Laigle et al., 2013).
175 Its northern part shows mainly extensional features (Bouysse & Guennoc, 1983; De Min
176 et al., 2015), including trench-perpendicular normal faults, from a latitude of $\sim 15^\circ$ all
177 the way up to the Anegada passage, which possibly represents the northernmost expres-
178 sion of this extensional system (Feuillet et al., 2002). South of 15° , the arc structure in-
179 cludes the Barbados accretionary prism extending up to 400 km eastward of the volcanic
180 island chain, bounded to the west by a well-developed 150-km-wide fore-arc basin. That
181 portion of the arc does not show the extensional structures observed in the north (Fig-
182 ure 1). The transition region between the north and south shows lateral ramps, follow-
183 ing the same trend as the Barracuda Ridge and the Tiburon Rise (e.g., Brown & West-
184 brook, 1987).

185 Current seismicity along the arc (Figure 1) shows that $M_W > 4$ events are mostly
186 focused in the northern part of the arc and around the El-Pilar-Central Range fault sys-
187 tem all the way in the south. In addition to seismicity highlighting the westward plunge
188 of the subducting slab, shallow seismicity occurs at crustal depths within the arc (i.e.,
189 ≤ 20 -40 km), particularly in the north. Seismicity is less prominent in the southern re-
190 gion that coincides with the sediment rich Barbados accretionary wedge. Schlaphorst et
191 al. (2016) analyzed seismicity patterns along the trench and found a high b-value (i.e.,
192 a higher fraction of small earthquakes) where fracture zones enter the trench. They did
193 not observe a clear difference in b-value distribution between the northern and south-
194 ern parts of the subduction zone. Bie et al. (2020) found a possible link between seis-
195 micity and fracture zones, with abundant intraslab seismicity beneath Martinique and
196 Dominica, where the Marathon and Mercurius fracture zones subduct. They also observe
197 pervasive seismicity in the cold mantle wedge corner, suggesting a deep decoupling depth
198 between the slab and the upper-plate mantle (Wada & Wang, 2009). This, in combina-
199 tion with the occurrence of the 2017 Martinique thrust event (M_W 5.8) at 51 km depth
200 suggests that the seismogenic zone may reach as far as ~ 65 km depth.

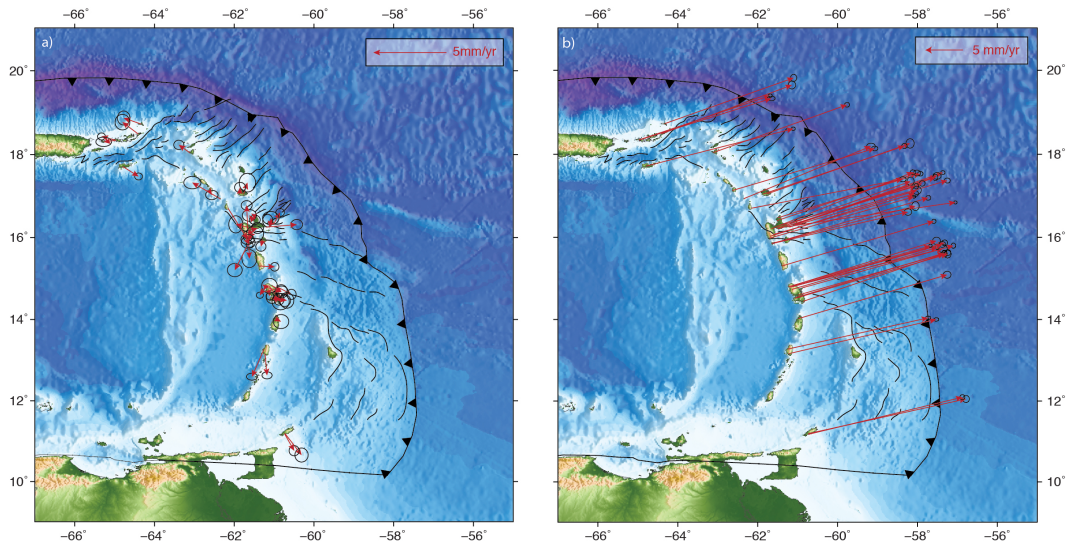


Figure 2. GPS velocities in the Caribbean (a) and North American (b) reference frames. Only velocities with uncertainties below 0.25 mm/yr are shown here for clarity, which is about 50% of the total dataset. Error ellipses are 95% confidence.

3 Methods

3.1 Geodetic Network and Data Processing

The GPS data used in this work come primarily from continuous GPS sites, some of them installed within the COCONet project (Braun et al., 2012). Additional campaign measurements are available on some of the islands, mostly in Martinique and Guadeloupe. The 74 velocities used here are a subset of the 445 stations that we routinely process that cover the entire Caribbean region. A list of all stations and GPS velocities is available in the supporting information. The data processing procedure is the same as used in Smithe et al. (2015) and is only briefly summarized hereafter.

We use the GAMIT-GLOBK software package (Herring et al., 2010) to process the double-difference phase measurements using the International Global Navigation Satellite Systems (GNSS) Service (IGS), Earth orientation parameters from the International Earth Rotation Service (IERS) products to produce loosely constrained daily solutions. We then combine these regional solutions with global daily solutions for the whole IGS network available from the Massachusetts Institute of Technology IGS Data Analysis Center into weekly position solutions. These weekly solutions are finally combined into a single position/velocity solution, which we tie to the International Terrestrial Reference Frame (ITRF2014, Altamimi et al., 2016) by minimizing position and velocity deviations from

219 a set of globally defined IGS reference sites common to our solution via a 12-parameter
220 Helmert transform.

221 At continuous GPS sites, we use the First-Order Gauss-Markov Extrapolation al-
222 gorithm (Herring, 2003; Reilinger et al., 2006) to obtain velocity uncertainties that ac-
223 count for time-correlated noise. For episodic sites, we include a $2 \text{ mm}/\sqrt{yr}$ random walk
224 component to account for colored noise in velocity uncertainties. Compared to the work
225 of Symithe et al. (2015), the solution used here contains at least 6 additional years of
226 data at the continuous sites. It also benefits from new GPS sites on some of the Lesser
227 Antilles islands.

228 In order to be able to solve for coupling on the Lesser Antilles subduction inter-
229 face, we rotate the velocities, originally expressed in ITRF, into a Caribbean-fixed ref-
230 erence frame. This operation is not trivial as (1) there are too few reliable GPS sites in
231 the interior of the – mostly oceanic – Caribbean plate to reliably estimate an angular
232 velocity, and (2) using an *a priori* angular velocity from other publications – even that
233 of Symithe et al. (2015) – would not insure consistency with our solution. We therefore
234 performed a Caribbean-wide kinematic inversion using the “blocks” code (Meade and
235 Loveless, 2009) following the same methodology and model geometry as in Symithe et
236 al.’s (2015) best-fit model. This procedure ensures an optimal definition of the Caribbean
237 frame as it uses a regional minimization that includes all sites in the solution, does not
238 require that we hand-select the sites that we *a priori* think belong to the Caribbean plate,
239 and is fully consistent with our velocity solution.

240 Figure 2 shows GPS velocities in both a Caribbean (a) and North American (b)
241 reference frame. Velocities in the Caribbean reference frame are very small, as found in
242 previous studies (López et al., 2006; Manaker et al., 2008; Symithe et al., 2015). A new
243 and intriguing aspect of this updated dataset is an apparent along-arc extension, as sites
244 in its northern part generally show NW-directed velocities (0.12-2.29 mm/yr) and sites
245 in its southern part show SSW-directed velocities (0.12-1.89 mm/yr). These residual ve-
246 locities appear significant at the 95% confidence interval at several of the continuous GPS
247 sites present in the solution. In the central part of the arc, from Martinique to Guade-
248 loupe, residual velocities in a Caribbean frame are more scattered but nonetheless show
249 a general ocean-ward direction, particularly consistent at sites in Guadeloupe and in the
250 eastern-most part of Guadeloupe. We do not observe a systematic pattern of west-directed
251 velocities, as one would expect if the plate interface was locked, even partially. This is
252 quantified in more details below.

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3.2 Inferring interseismic coupling

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To model the interseismic coupling along the subduction interface, we invert the GPS observations using a Bayesian approach and a realistic geometry of the plate interface. Previous studies that used GPS velocities to estimate coupling used a planar subduction geometry, with a constant dip angle of 16° (e.g., Symithe et al., 2015). Since then, more detailed models of the subduction interface have become available, allowing us to better account for the influence of fault geometry in the inversion process. Here we test two different fault geometries: the Slab2 model (Hayes et al., 2018) and a more recent model developed by Bie et al. (2020). We discretize the subduction interface into triangular elements, which vary in size from 2500 km^2 (i.e., $\sim 70 \text{ km}$ side-length) below the islands, to 11500 km^2 (i.e., $\sim 150 \text{ km}$ side-length) along the shallow parts of the fault. The size variability allows us to account for the increasing distance and hence decreasing model sensitivity between the fault and the island arc as one goes towards the trench.

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We adopt a backslip (slip deficit) approach to estimate interseismic coupling from geodetic displacement rates, in which deformation related to interseismic locking along the subduction interface is modeled by continuous slip of the locked part in a reverse sense compared to coseismic slip (Savage, 1983). We model the measured GPS velocities as the result of both interseismic coupling along the subduction interface and homogeneously distributed strain within the arc. The Green's functions that relate slip along the fault to displacement at the surface, are calculated using a layered semi-infinite elastic medium (Zhu & Rivera, 2002). We implement a crustal structure based on the four-layer velocity model proposed by Schlaphorst et al. (2018), who used receiver function inversions to obtain 1D velocity profiles for all islands along the Lesser Antilles arc (Figure 3). Based on the range of velocities they propose for each layer, as well as the velocities proposed by other models (Bie et al., 2020; Raffaele, 2012), we assume a 15% uncertainty on the elastic parameters defining the overall crustal structure.

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The relation between the data measured at the surface and modelled slip along the fault can be described by the forward problem $\mathbf{d} = \mathbf{G}\mathbf{m}$, where \mathbf{d} represents the data vector containing the horizontal GPS velocities measured at the islands, \mathbf{G} the Green's functions matrix (i.e., the matrix relating interseismic coupling to surface displacements) and \mathbf{m} the vector of model parameters (i.e., the vector containing values of fault coupling for each fault element). The goal is to infer the distribution of model parameters (\mathbf{m}) that is consistent with our data (\mathbf{d}). Because of data and model uncertainties and the uneven distribution of GPS sites at the surface, the solution to such an inverse problem is non-unique. Therefore, model uncertainties estimated in a least-squares sense for the 'best-fit' solution provide limited information on the actual quality of the fit of the

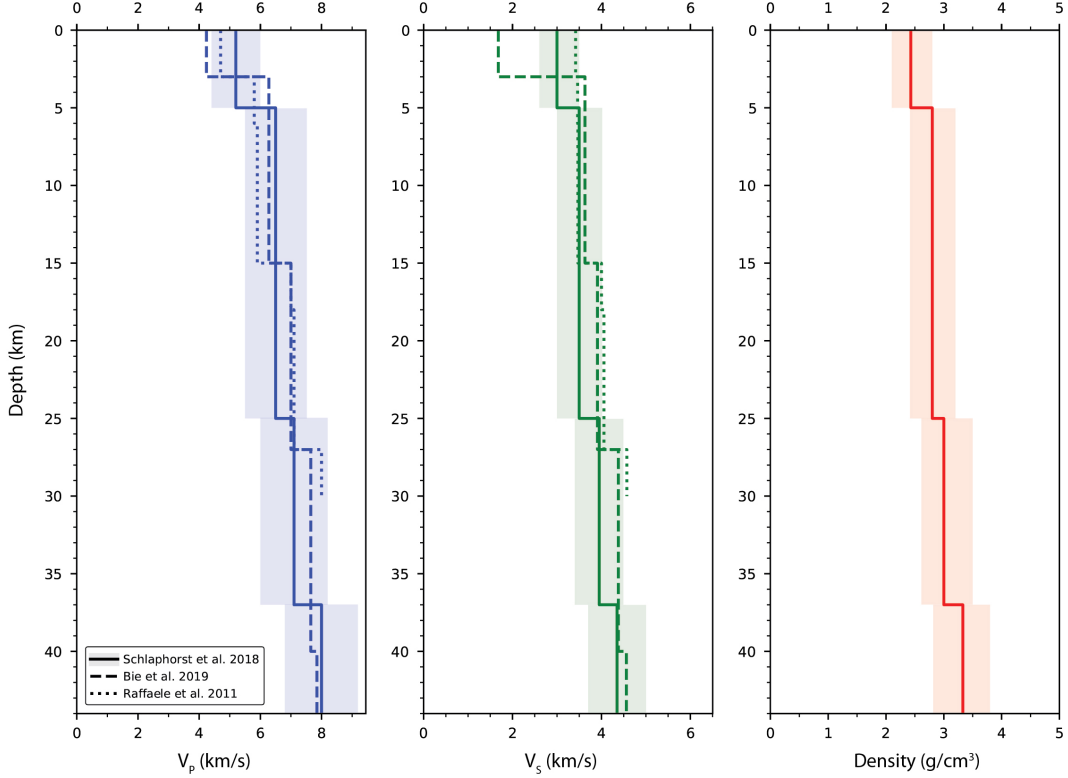


Figure 3. Crustal structure showing P-wave velocity (V_P), S-wave velocity (V_S) and density as a function of depth. The solid lines represent the model used in this study (based on Schlaphorst et al. (2018)). Shaded areas represent a 15% uncertainty region. Several other models are indicated by the dashed lines.

289 data to the model. Instead of deriving a single solution of interseismic coupling, we adopt
 290 a Bayesian approach that explores the entire range of possible models and provides a prob-
 291 abilistic estimate of interseismic coupling (Minson et al., 2013). These estimates do not
 292 rely on any spatial smoothing and include a realistic approximation of uncertainties re-
 293 lated to measurement- and modeling errors. The ensemble of plausible models that fit
 294 the observations and are consistent with prior constraints are described by full poste-
 295 rior probability distributions. Such a probabilistic approach allows us to objectively as-
 296 sess the whole range of model parameters allowed by the data. Following Bayes' theo-
 297 rem, we write the posterior probability density function (hereafter PDF) of the model,
 298 $p(\mathbf{m}|\mathbf{d})$, as,

$$p(\mathbf{m}|\mathbf{d}) \propto p(\mathbf{m}) \exp\left[-\frac{1}{2}(\mathbf{d} - \mathbf{G}\mathbf{m})^T \mathbf{C}_\chi^{-1} (\mathbf{d} - \mathbf{G}\mathbf{m})\right] \quad (1)$$

299 where $p(\mathbf{m})$ represents the prior PDF of the model and \mathbf{C}_χ the misfit covariance matrix
 300 in the data space. The prior PDF describes our state of knowledge before considering

301 the data. Here we use a uniform (box-like) prior between 0 (i.e., the megathrust slips
 302 at plate convergence rate) and 1 (i.e. the megathrust is locked). We therefore assume
 303 no prior knowledge on the model parameters and an equal likelihood for all possible val-
 304 ues of interseismic coupling.

305 The misfit covariance matrix \mathbf{C}_χ represents the sum of the data covariance matrix
 306 \mathbf{C}_d , describing the uncertainties on the data, \mathbf{d} , and the prediction error matrix, \mathbf{C}_p , which
 307 describes uncertainties of the model predictions such that:

$$\mathbf{C}_\chi = \mathbf{C}_d + \mathbf{C}_p \quad (2)$$

308 The quality of the model predictions, \mathbf{C}_p , is mainly influenced by the imperfect knowl-
 309 edge of the Earth structure (i.e., the elastic parameters V_p , V_s and ρ). In order not to
 310 overfit the data and produce reasonable estimates of coupling uncertainties along the fault,
 311 we need a careful description of the errors. For this, we use a stochastic forward model
 312 developed by Duputel et al. (2014), based on a linear formulation of the prediction un-
 313 certainty. Rather than providing a single set of predictions for a given source model, as
 314 would be done in a deterministic approach, this stochastic formulation produces a dis-
 315 tribution of predictions for a given uncertainty in the elastic structure (i.e., 15%, as in-
 316 dicated above).

317 Since we are dealing with a high-dimensional model space, the solution of our in-
 318 verse problem cannot be characterized using analytical techniques or simple Metropolis-
 319 like sampling. We therefore explore the model space in a random manner, sampling the
 320 posterior PDF, $p(\mathbf{m}|\mathbf{d})$, using AlTar, a parallel Markov Chain Monte Carlo (MCMC) al-
 321 gorithm based on the Cascading Adaptive Transitional Metropolis in Parallel (CATMIP)
 322 algorithm (Minson et al., 2013). The MCMC method uses a random walk to explore the
 323 model space and probabilistically determines whether to take a certain step or not. Al-
 324 Tar runs thousands of these MCMC chains in parallel, in order to efficiently and exhaus-
 325 tively sample the model space. Rather than sampling the posterior PDF immediately,
 326 a transitioning approach is used, thereby first sampling the prior PDF, $p(\mathbf{m})$, and then
 327 slowly increasing the information brought by the data until the posterior PDF is sam-
 328 pled. Computational tractability is ensured via the use of multiple Graphics Processing
 329 Units (GPUs) in parallel.

330 Finally, we end up with an ensemble of 150,000 models drawn from the posterior
 331 PDF. From these models, we can explore various statistical properties, such as the mean,
 332 mode, standard deviation, kurtosis, skewness, and information gain (i.e., with respect
 333 to the prior). In addition, we can explore the probability densities for each fault element
 334 individually (i.e., the marginal PDF's).

4 Results

In the following sections, results from several analyses regarding the interseismic coupling along the subduction megathrust will be discussed. We present the model sensitivity (section 4.1.), some simple forward models to understand what our model would predict for various coupling scenarios (section 4.2.), the posterior PDF resulting from the Bayesian inversion (section 4.3.), a comparison between two different slab geometry models (section 4.4.) and some specific tests regarding the historical 1839 and 1843 earthquakes (section 4.5.). Except for section 4.4., all results are based on the slab geometry from the Slab2 model (Hayes et al., 2018), although section 4.4. will demonstrate that similar results would be observed when using the slightly steeper slab geometry proposed recently by Bie et al. (2020).

4.1 Model Sensitivity

In order to evaluate the robustness of the fault coupling estimates, we compute the sensitivity, \mathbf{S} , of the model to the GPS dataset, defined as,

$$\mathbf{S} = \text{diag}(\mathbf{G}^t \mathbf{G}) \quad (3)$$

where \mathbf{G} is the Green's functions matrix defined previously and *diag* is the diagonal operator that extracts the diagonal after multiplication (Loveless & Meade, 2011; Lin et al., 2015). For each node of the fault, \mathbf{S} describes the sum of squared displacements at all data locations resulting from a coupling of 1 on that specific node. The sensitivity therefore indicates the relative contribution of each node to the prediction of surface displacements. It provides a useful estimate of the extent to where the data is able to inform the posterior PDF of the model, and where it will hence differ the most from the uniform prior PDF. Nodes located further away from data locations are generally expected to have lower sensitivity and will usually have larger uncertainties in the posterior PDF.

Figure 4 shows the model sensitivity, based on the Slab2 geometry (Hayes et al., 2018) and the elastic structure presented previously (Schlaphorst et al., 2018). We observe a higher sensitivity for the central part of the seismogenic zone, between 25 km and 60 km depth. The region surrounding Guadeloupe has the highest sensitivity, extending even down to 100 km depth, the downdip limit of our fault model. As expected, we find the lowest sensitivity closest to the trench, as these nodes are the furthest away from the data locations. This is particularly the case for the southern part of the subduction zone, where the slab dip is shallower (i.e., $\sim 7^\circ$ with respect to $\sim 14^\circ$ in the north) and the trench is located ~ 200 km further to the East with respect to the islands. We note

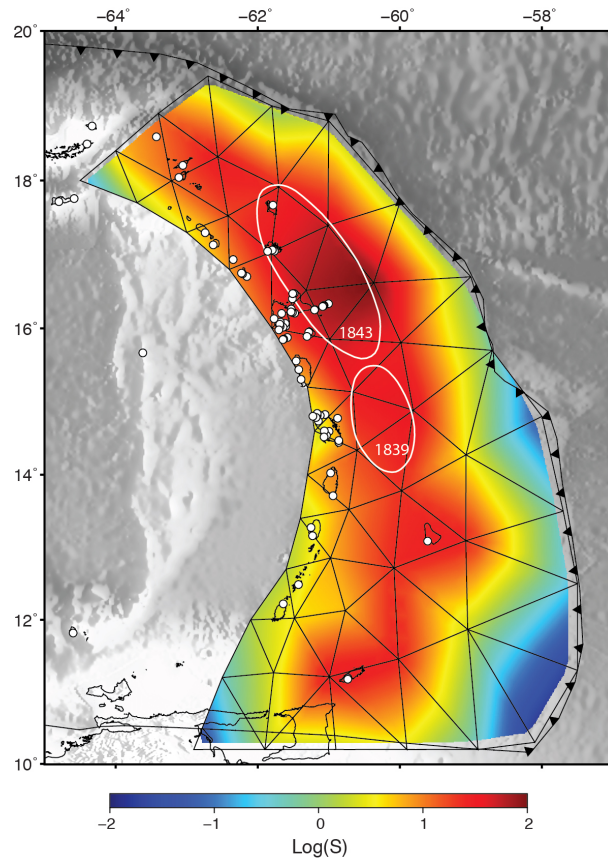


Figure 4. Model sensitivity (based on the Slab2 geometry), describing how well the GPS stations on the islands (white dots) can constrain the plate interface behaviour. Each node is colored by the sum of the displacement at the GPS stations, due to unit coupling along that node.

368 that the areas in which the 1839 and 1843 earthquakes are thought to have occurred cor-
 369 respond to the highest sensitivity areas of the fault model. The model sensitivity for the
 370 Bie et al. (2020) slab geometry and some tests that explore different distributions of data
 371 locations are included in the supporting information (Figures S1-S3).

372 4.2 Simple forward models

373 We start our search for a better understanding of the distribution of interseismic
 374 coupling by manually exploring forward models in order to develop an intuition on the
 375 velocity magnitudes to expect for certain scenarios of interplate coupling. We test an in-
 376 terface that is either homogeneously or partially (i.e., 50%) locked down to a depth of
 377 1) 20 km, representing the shallowest part of the subduction interface or 2) 65 km, cur-

378 rently believed to represent a minimum downdip limit of the seismogenic zone (Bie et
 379 al., 2020).

380 Figure 5 shows forward models and the resulting predictions. The model with full
 381 coupling down to 20 km, predicts westward velocities of 1-2 mm/yr at the islands that
 382 are closest to the trench (i.e., Barbados in the south, and Barbuda, St Martin and An-
 383 guilla in the north). The real observations on these islands are similar in magnitude, but
 384 are oriented in a trench-parallel direction (i.e., towards the northwest and south) rather
 385 than trench-perpendicular as the response to interplate coupling shows. This indicates
 386 that despite the relatively low sensitivity for these shallow parts of the plate interface,
 387 a fully-coupled interface down to 20 km depth would be detected by the stations on the
 388 above-mentioned islands. This is less clear however, for 50% coupling within this depth
 389 range. In the case of an interface coupled down to 65 km depth, the synthetics are clearly
 390 inconsistent with the observed GPS velocities, both for the fully- and partially coupled
 391 scenarios. With an interface that is fully coupled, the synthetics indicate westward ve-
 392 locities at all stations, that reach up to 15.07 mm/yr, about 7 times larger than obser-
 393 vations.

394 These forward models indicate that both a partially- and a homogeneously-locked
 395 interface down to 65 km depth are very unlikely. A fully locked interface down to 20 km
 396 also seems unlikely, due to the difference in orientation between data and predictions for
 397 islands closer to the trench. However, a partial (less than 40%) locking along the shal-
 398 low parts of the megathrust cannot be excluded based on these first tests.

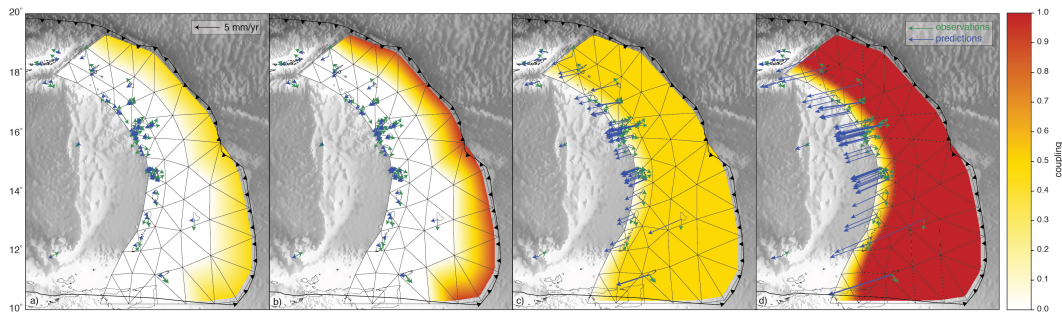


Figure 5. Forward models showing synthetic velocities (blue arrows) as a result of different locking scenarios in comparison with measured GPS velocities (green arrows). The fault is either locked down to 20 km depth (a and b), or 65 km depth (c and d). The different colors indicate two models of coupling: fully locked (i.e., coupling = 1.0) and 50% locked (i.e., coupling = 0.5).

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4.3 Posterior Interseismic Coupling Distribution

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We will now discuss the results of the Bayesian inversion, where Figure 6a shows the distribution of coupling corresponding to the mean of the posterior PDF. In general, the inferred coupling is very low (< 0.2), especially in the central parts of the seismogenic zone, where we also observe the highest sensitivity (Figure 4). Along the shallower parts of the interface, we find a mean coupling of around 0.2, while along the deeper parts (i.e. > 60 km) mean coupling varies between zero around the islands of St Kitts & Nevis, to 0.5 below Martinique and a local high of 0.7 west of the Grenadines. Figure 6b shows the mode of the posterior PDF, highlighting the most common values of coupling derived from the marginal PDF for each node. It shows zero coupling everywhere, except for two local highs along the deeper part of the subduction interface (i.e., 60-100 km), one below Martinique (coupling of 0.5), and one west of the Grenadines (coupling of 0.8).

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Both the mean and mode of the posterior PDF only provide part of the information on the estimated interseismic coupling, as one also needs to consider the width of the distribution for each node and how much the posterior PDF has evolved from a uniform prior with a mean coupling of 0.5. This can be better understood by looking at the marginal PDFs for each individual node (nodes 1-5 in figure 6a). Nodes 1,2 and 5 show distributions with the highest probability around a coupling of 0, with an especially narrow distribution for node 2, located in the region with highest sensitivity (Figure 4). Node 3 shows a wide PDF centered around 0.5, meaning that it has evolved the least from the uniform (box-like) prior PDF. Node 4 shows a PDF with a peak near a coupling value of 1, while surrounding nodes have their highest probability concentrated around 0 again. Because the depth of both nodes 3 and 4 (i.e., 100 km) places them below the downdip limit of the seismogenic zone, we interpret these values as outliers along a generally uncoupled interface. They could be a consequence of the model trying to best fit some of the southward GPS velocities on the islands. Figure S4 in the supporting information confirms this by showing southward oriented surface predictions related to a forward model where only these two nodes are fully locked. A comparison between GPS observations and model predictions based on the mean posterior PDF can be found in Figure 7a. They generally agree well in terms of velocity magnitude, though not always in direction. It is however difficult to compare velocities that are in the 0.2-2 mm/yr range with 95% confidence uncertainties that are often close to the observed signal.

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In order to account for the along-arc extension pattern observed in the GPS velocities described above, we jointly solve for homogeneously distributed surface strain together with the plate interface coupling. We estimate a single horizontal strain rate tensor (i.e., 3 unknowns) for the whole arc in order to limit the number of parameters

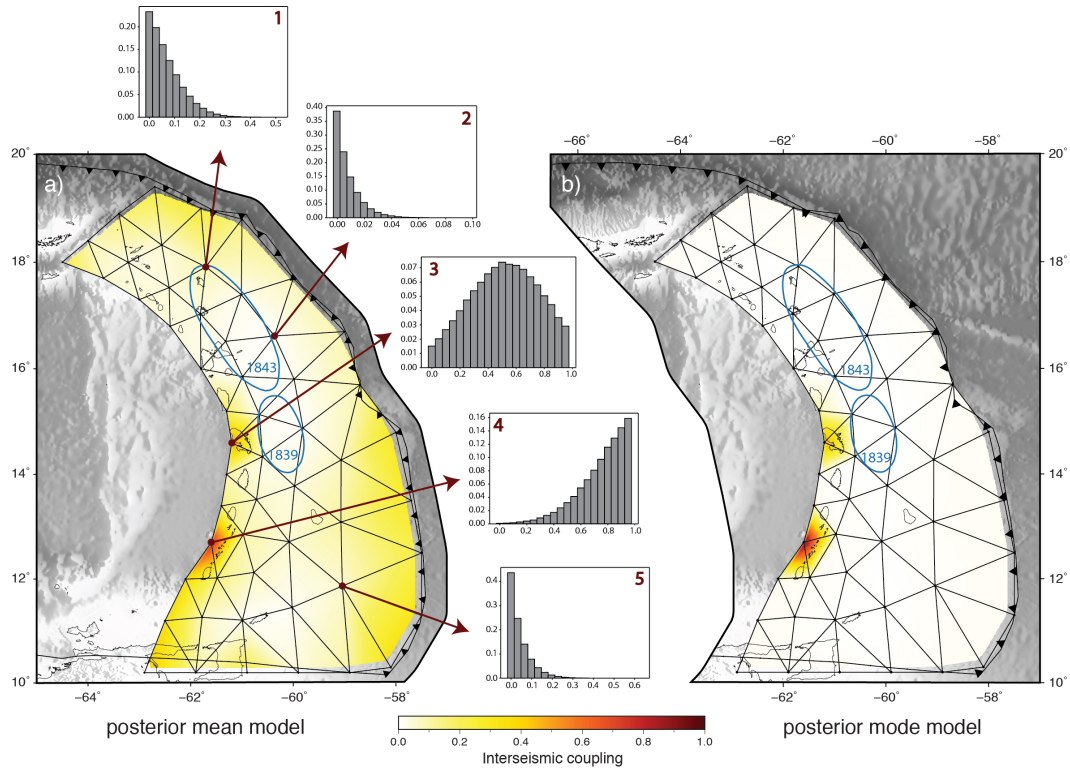


Figure 6. Posterior mean (a) and mode (b) coupling models for the Slab2 geometry. The inversion provides probability density distributions for each node of the triangular mesh, of which the mean and mode values are shown in the two maps. The marginal probability densities for several nodes are shown as well. The blue contours indicate the proposed rupture contours of the historical 1839 and 1843 earthquakes.

435 to be inverted for. Figure 7b shows the result of this estimation in terms of model ve-
 436 locities at the GPS sites. Although these velocities are quite low (i.e., ~ 0.03 to 0.70 mm/year),
 437 a clear pattern of north-south extension emerges. This indicates that the GPS data do
 438 contain the extension observed geologically along the arc (Bouysse & Guennoc, 1983; Feuillet
 439 et al., 2002; De Min et al., 2015; Münch et al., 2014) and can now provide a quan-
 440 titative estimate of the slip rate on intra-arc normal faults. The results from this inver-
 441 sion indicate that the total amount of fault slip is unlikely to exceed 1 mm/yr, though
 442 a proper estimate would require discretizing the strain rate estimation. We are however
 443 limited by the number and location of islands.

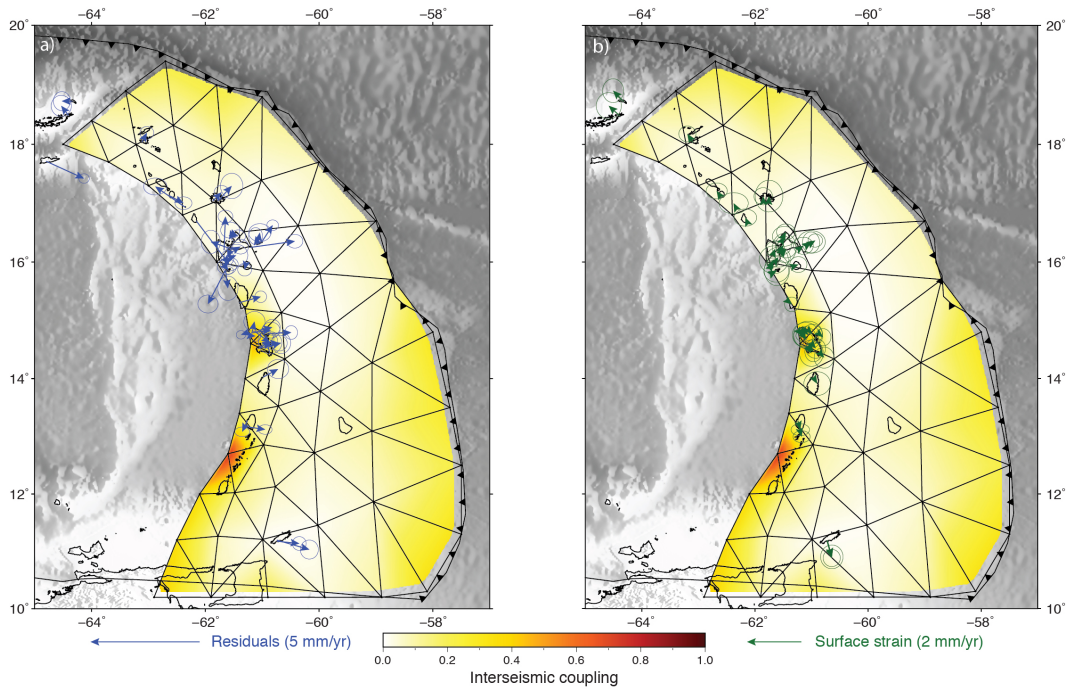


Figure 7. Posterior mean model for the Slab2 geometry, with residual velocities (a) and extension estimated by the model, based on a homogeneous strain tensor (b). As in Figure 2, only velocities with uncertainties below 0.25 mm/yr are shown here for clarity, which is about 50% of the total dataset. Error ellipses are 95% confidence.

444 4.4 The role of Slab Geometry

445 Previous studies that attempted to infer interseismic coupling along the Lesser An-
 446 tilles subduction interface used a planar and constant fault geometry (e.g. 16 °, Symithe
 447 et al., 2015). Uncertainties in subduction interface geometry are a limitation to our abil-
 448 ity to accurately estimate interplate coupling (Paulatto et al., 2017). The recent, more
 449 detailed subduction interface models proposed by Hayes et al. (2018) and Bie et al. (2020),
 450 allow us to test how a change in fault geometry affects the posterior PDF of interseis-
 451 mic coupling inferred from the GPS data. This could be assessed from the posterior PDF
 452 using the approach of Ragon et al. (2018), but we prefer to directly show the difference
 453 between two models with two plausible geometries rather than lumping this effect within
 454 the posterior PDF.

455 Figure 8 shows the mean and mode posterior coupling estimates for both geome-
 456 tries, as well as three depth profiles along sections of the arc, in order to highlight the
 457 differences in slab geometries. The geometry proposed by Bie et al. (2020) fits the lo-
 458 cal seismicity (i.e., the CDSA catalog) better and might therefore better represent the

459 actual geometry of the Lesser Antilles slab. We however find that the difference of mean
 460 interseismic coupling between the two geometries is very small and that the two mod-
 461 els are in very good agreement. Both models show very low to low coupling along most
 462 parts of the interface, except for the two local highs discussed previously. We observe
 463 slightly larger uncertainties in the model based on the Bie et al. (2020) geometry in the
 464 regions where this model becomes steeper than the Slab2 model and is therefore located
 465 further away from the GPS observations.

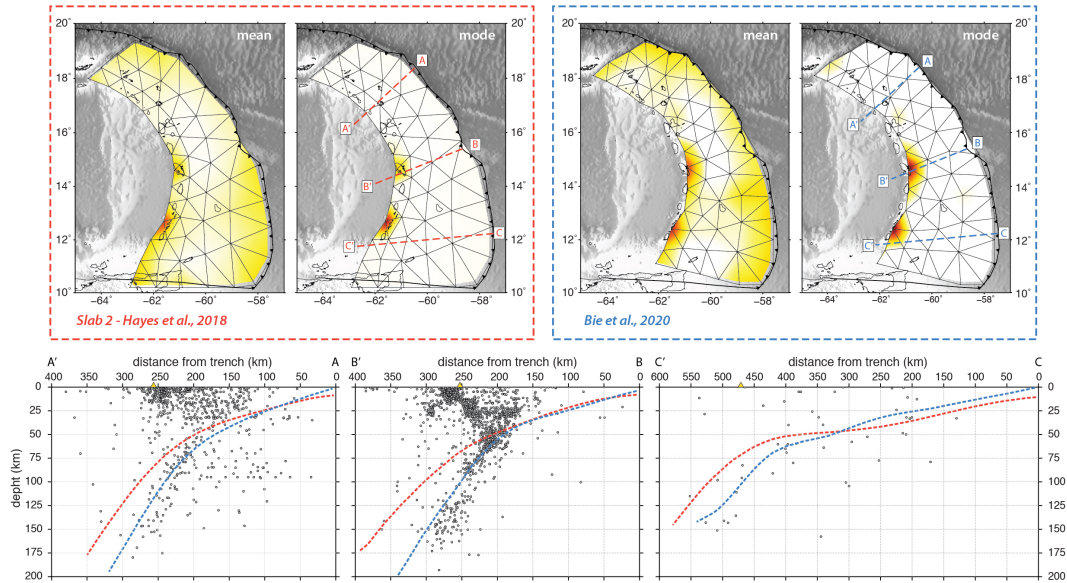


Figure 8. Posterior mean and mode coupling for the Slab2 geometry vs. the geometry proposed by Bie et al. (2020). Three depth profiles are indicated in the maps as dashed colored lines: red for the Slab2 geometry and blue for the Bie et al. (2020) geometry. Seismicity from the CDSA catalog (1972-2013) is plotted in grey. The yellow triangles indicate the locations where the profile intersects the volcanic arc. For a colorscale of interseismic coupling, see Figure 6 and 7.

466 4.5 A Test of the the 1843 and 1839 Earthquake Sources

467 The overall low coupling found along the Lesser Antilles subduction interface in
 468 this study raises questions about the location and faulting mechanism of the historical
 469 1839 and 1843 earthquakes. Assuming these events were thrusts along the subduction
 470 interface, current earthquake cycle models (e.g., Avouac, 2015) predict that the rupture
 471 areas should have healed and re-locked. In order to test whether we would detect such
 472 re-locking, we calculate the predicted velocities as a result of full locking of the proposed

473 1839 and 1843 rupture areas (Feuillet et al., 2011). For this, a refined mesh was used to
 474 accurately lock the plate interface segments associated with the 1839 and 1843 events.
 475 We then use the synthetic velocities, with the uncertainties of the observed data set, as
 476 input to the inverse model described above (section 3.2). For this inversion stage, we use
 477 the same fault discretization as used before (section 4.3), meaning that we cannot re-
 478 trieve the same coupling pattern as was imposed in the forward model (i.e., with a lo-
 479 cally refined mesh). The results with a similar fault discretization for both the forward
 480 model and inversion can be found in the supporting information (Figure S5).

481 Figure 9 shows the result of this forward model and inversion for the 1839 and 1843
 482 events. Both the mean and mode posterior coupling estimates retrieve the coupling we
 483 imposed in the forward models. The areas updip of these locked regions also show some
 484 degrees of coupling, likely related to the lower sensitivity and therefore the reduced ca-
 485 pacity of our model to correctly infer coupling in these distal regions. Overall, these re-
 486 sults indicate that if the 1839 and 1843 rupture areas had re-locked, they should (1) in-
 487 duce westward interseismic velocities of up to 7 mm/yr in Guadeloupe and Martinique
 488 that we do not observe in the GPS data, and (2) be detected as locked patches in the
 489 inverse models described above (section 4.3). Since these central regions of the plate in-
 490 terface also have the highest sensitivity to the GPS data (Figure 4), we argue that it is
 491 unlikely that they have re-locked.

492 5 Discussion

493 The sensitivity and forward model results (Figures 4 and 5) demonstrate that a to-
 494 tal or partial locking of the subduction interface in the 20-65 km depth range would in-
 495 duce a plate boundary deformation signal with detectable, Caribbean-ward, velocities
 496 at the GPS sites on the Lesser Antilles islands. This is especially true in the northern
 497 part of the arc, where trench-to-island distances, ranging from 160 to 250 km, are sim-
 498 ilar to Japan or South America, where strain accumulation as a consequence of a locked
 499 interface is recorded by the coastal GPS stations (e.g., Loveless & Meade, 2010; Maz-
 500 zotti et al., 2000; Nocquet et al., 2014). In the southern part of the Lesser Antilles arc,
 501 where slab dip decreases, the increasing trench-to-island distance is reflected in the re-
 502 duced model sensitivity close to the trench. Forward and Bayesian inverse models, as
 503 well as the specific tests for the 1839 and 1843 events all show that the Lesser Antilles
 504 subduction interface currently has low to very low coupling. As a result, the active plate
 505 margin is unlikely to be accumulating elastic strain at a significant rate today. This low
 506 interplate coupling and low elastic strain accumulation rate raise questions about the na-

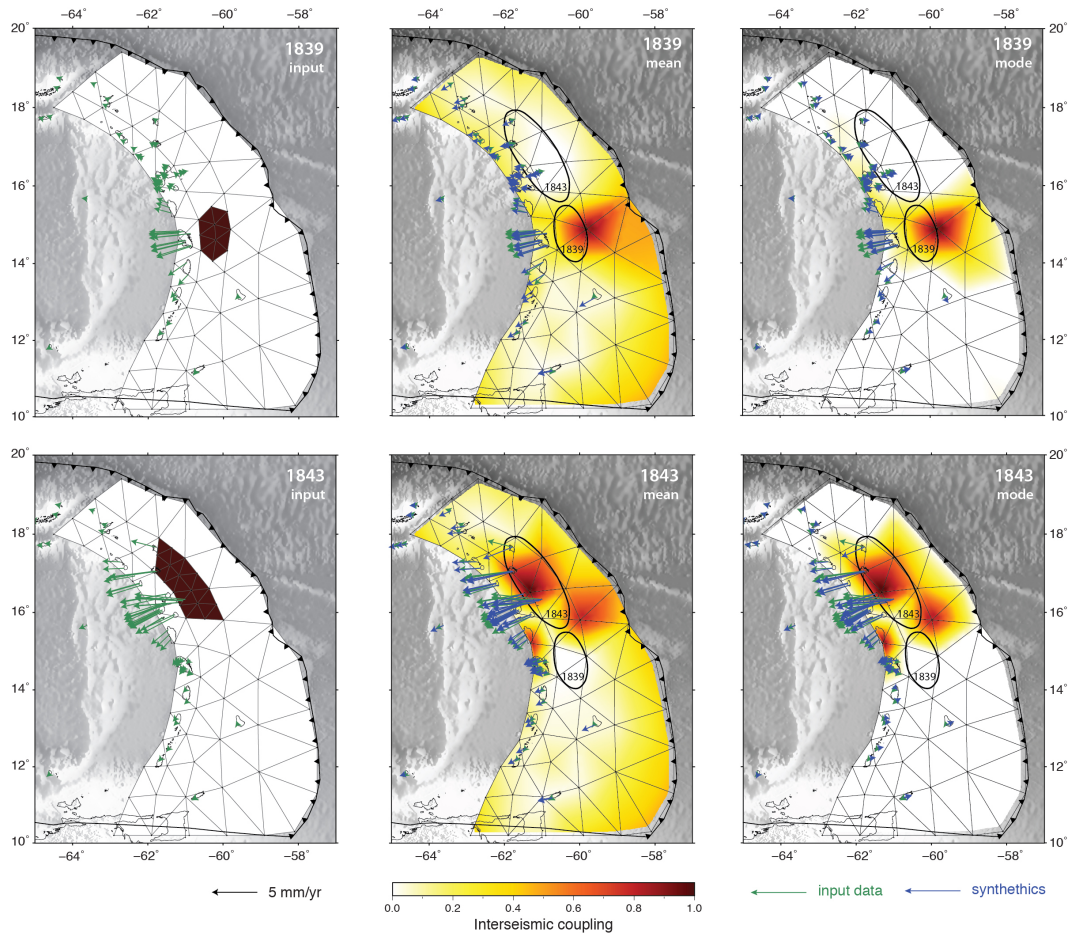


Figure 9. Forward model + inversion for the proposed 1839 and 1843 rupture areas. The panels show (from left to right) the forward models with the resulting GPS velocities (green arrows), and the mean and mode of the posterior PDF. The synthetic velocities resulting from inferred coupling models are indicated with blue arrows.

507 ture of the 1839 and 1843 earthquakes, as well as the physical mechanism that is respon-
 508 sible for the low coupling we observe.

509 **5.1 What is the nature of the 1839 and 1843 events?**

510 Because of their magnitude and location at a subduction plate boundary, the 1839
 511 and 1843 earthquakes are often considered as thrust events on the plate interface (Bernard
 512 & Lambert, 1988; McCann & Sykes, 1984). However, no direct evidence for this exists
 513 yet, and the magnitude and location of these historical events remain debated. Magni-
 514 tude estimates for the 1843 event are mainly based on reported intensities. Early esti-
 515 mates range from 7.5 to 8.5, with estimated rupture lengths ranging from 100 to 300 km

516 (Bernard & Lambert, 1988; ten Brink et al., 2011; Feuillet et al., 2011; Hough, 2013).
 517 By including additional felt reports from the east coast of the United States, Hough (2013)
 518 proposed a magnitude of M_W 8.4 with values as high as M_W 8.5-8.7 if the earthquake
 519 occurred farther offshore than its generally preferred location beneath the islands of Guade-
 520 loupe.

521 The absence of a tsunami or noticeable vertical deformation of the coasts of Guade-
 522 loupe or Antigua (Bernard & Lambert, 1988) is conspicuous, since all $M_W \geq 8.4$ sub-
 523 duction megathrust event in the instrumental record resulted in a tsunami (National Geo-
 524 physical Data Center / World Data Service, 2020), with maximum water heights rang-
 525 ing from 4.2 m to 42 m (for events between M_W 8.4 to 8.7). As the rupture extent of
 526 past earthquakes is being re-visited (Lay & Rhode, 2019; Sladen & Trevisan, 2018), it
 527 is becoming more and more clear that such great events often include near-surface rup-
 528 tures. They generally saturate the downdip width of the seismogenic zone, before prop-
 529 agating hundreds of kilometers along-strike (Heuret et al., 2011). A rupture of the deeper
 530 part of the plate interface only, as typically proposed for the 1843 event, is yet to be ob-
 531 served in great ($\sim M_W 8.4$) megathrust earthquakes of the instrumental time period.
 532 This all, in combination with the very low coupling of the proposed rupture area found
 533 in this study, suggests that the 1843 event either had a smaller magnitude, or had a dif-
 534 ferent faulting mechanism, and therefore did not occur along the subduction interface.

535 The instrumental record shows that all $M > 7$ earthquakes of the Lesser Antilles in
 536 the past ~ 70 years have a normal faulting mechanism: M 7.5 in 1953, M_S 7.5 in 1969,
 537 M_S 7.4 in 1974 and M_W 7.4 in 2007. Both the M_W 7.4 2007 Martinique event and the
 538 M 7.5 1953 St. Lucia events have been interpreted as intraslab normal faulting events,
 539 that occurred at depths of 156 and 135 km, respectively. A similar mechanism and depth
 540 are plausible for the 1839 event, which as has similar magnitude and occurred in the same
 541 region, characterized by dense intermediate-depth seismicity. Intra-slab, normal fault-
 542 ing, and intermediate-depth earthquakes as large as M_W 8.5, if this was indeed the mag-
 543 nitude of the 1843 event, have however not been observed in the instrumental record.
 544 The 2019, M_W 8.1 intraslab and normal-faulting event at 110 km in Peru shows how-
 545 ever that larger events can also occur at intermediate depths. Though of smaller mag-
 546 nitude than some of the estimates for the 1843, Lesser Antilles, earthquake, this event
 547 was felt all over South America (Jiménez et al., 2020), with macroseismic intensities at
 548 large distances that are similar to those reported for the 1843 Lesser Antilles event (e.g.,
 549 IV MMI at ~ 700 km distance, Hough, 2013; Jiménez et al., 2020). A large, intermediate-
 550 depth rupture would also explain the large felt extent of the 1843 event and the absence
 551 of a noticeable tsunami.

5.2 What physical mechanism is responsible for the low coupling?

The subduction of topographic features has been proposed to play a role in tuning lateral variations of plate coupling and therefore mega-earthquake occurrence (e.g., Lallemand et al., 2018; Wang & Bilek, 2014). In the Lesser Antilles, the subduction of fracture zones, or of oceanic ridges like the Barracuda Ridge and Tiburon Rise, has long been proposed to segment the seismogenic zone (McCann & Sykes, 1984). More recent studies found that larger b -values, indicative of stress release through a higher fraction of small earthquakes, and low shear-wave velocities correlate with the location of incoming fracture zones on the American plates (Cooper et al., 2020; Schlaphorst et al., 2016). They relate this to excess dehydration due to fluids that are delivered into the subduction by the fracture zones. Such fluids along the plate interface will allow rupture at lower stress levels due to higher pore fluid pressures and hence increase the number of small earthquakes. Following that hypothesis, these incoming fracture zones and ridges facilitate stress dissipation through aseismic processes and should then act as "low coupling" areas. Such low coupling areas would then act as barriers to the propagation of megathrust earthquakes, hence limiting their magnitude. This assumption of a seismogenic segmentation by the incoming Tiburion Rise and Baracuda and Saint-Lucia ridges was also made by Hayes et al. (2014) to quantify the earthquake and tsunami potential of the Lesser Antilles subduction. However, the inversion of GPS velocities described above does not show variations in interseismic coupling that correlate with the presence of subducting ridges or fracture zones. Furthermore, since we find homogeneous low coupling along the entire subduction interface, it is unlikely that localized features play a dominant role here.

Another characteristic of the Lesser Antilles, that holds for the entire region, including Puerto Rico, is the subduction of slow-spread oceanic lithosphere formed along the Mid Atlantic and Proto-Caribbean Ridges. Slow-spreading ridges create an oceanic lithosphere that is more heterogeneous in terms of thickness and composition, and more pervasively hydrated than their fast-spreading counterparts (Paulatto et al., 2017). As this hydrated oceanic lithosphere subducts, dehydration metamorphic reactions release fluids that migrate upwards, which could explain the high V_p/V_s ratios (i.e., a proxy for high pore-fluid pressure) in the central part of the Lesser Antilles forearc (Martinique - Antigua; Paulatto et al., 2017). Increased pore-fluid pressures along the subduction interface reduce the effective normal stress and may therefore promote stable creep (Audet & Schwartz, 2013; Bilek & Lay, 2018; Moreno et al., 2014; Saffer & Tobin, 2011). A negative correlation between interplate coupling and high V_p/V_s ratios has indeed been observed before (Moreno et al., 2014), as well as a positive correlation between the amount of subducting fluids and the occurrence of intermediate-depth earthquakes (Faccenda et

588 al., 2012; Hacker et al., 2003). A hydrated oceanic crust has also been associated with
589 creep along the deeper parts of the subduction interface (i.e., in the 370° to 450° tem-
590 perature range), because a weak phyllosilicate-bearing mineralogy may allow the crust
591 to creep at shear stresses low enough to accommodate significant plate interface displace-
592 ment (Tulley et al., 2020). The subduction of fluid-rich slow-spread lithosphere is there-
593 fore an important candidate to explain the low coupling of the Lesser Antilles subduc-
594 tion inferred from GPS observations.

595 Looking at global subduction zones and seismogenic behaviour, several other re-
596 gions are thought to be mainly aseismic, such as the Aegean, Calabria, South Sandwich
597 and Mariana subduction zones (e.g., Carafa et al., 2018; Ruff & Kanamori, 1983; Vanneste
598 & Larter, 2002; Vernant et al., 2014). What these regions all have in common are their
599 short length and strong curvature. In a global comparison of geometric subduction zone
600 parameters with maximum megathrust earthquake magnitude, Schellart and Rawlinson
601 (2013) found that stronger trench curvature correlates with fewer great megathrust earth-
602 quakes. The physical reason invoked is that rupture propagation over long distances is
603 favored by a relatively planar subduction interface, but hindered by curved segments in
604 subduction zones.

605 The lesser Antilles and Mariana subduction zones also share evidence for trench-
606 parallel extension in the form of arc-perpendicular normal faults (Feuillet et al., 2002;
607 Stern & Smoot, 1998). In the Lesser Antilles, we are now able to document, from GPS
608 observations, that this extension concerns the entire arc (Figure 7). In addition, the east-
609 ward (i.e., ocean-ward) GPS velocities observed in the central part of the Lesser Antilles
610 arc show that an additional trench-perpendicular component of extension exists. In the
611 Calabrian and Aegean subduction zones, forearc extension has been documented as well
612 (Caputo et al., 2010; D’Agostino et al., 2011; Marsellos et al., 2010; Totaro et al., 2016),
613 suggesting a possible link to the aseismic character of these four subduction zones. Ex-
614 tension in the overriding plate has been proposed to play a role in controlling the downdip
615 limit of interseismic coupling (Wallace et al., 2012), and could therefore also be impor-
616 tant in tuning the overall seismogenic behaviour of a margin.

617 **6 Conclusions**

618 We provide a new assessment of interseismic coupling for the Lesser Antilles sub-
619 duction zone, based on updated GPS velocities and the latest models of the slab geom-
620 etry and elastic crustal structure. We use a Bayesian approach, allowing us to explore
621 the entire range of plausible models and to provide realistic estimates of the state of cou-
622 pling along the subduction interface. We find low to very low coupling along the entire

623 plate interface, including in the proposed rupture areas of the 1839 and 1843 earthquakes.
624 Given the fact that already ~ 175 years have passed since the 1843 event, following the
625 reasoning of current earthquake cycle models (e.g., Avouac, 2015; Savage, 1983), at least
626 a partial re-locking of some regions would be expected in the case of a large megathrust
627 event. This all questions the notion that these historical earthquakes were thrust events
628 on the plate interface. While a further understanding of temporal variations in interseis-
629 mic coupling needs to be addressed by future geodetic and geologic observations, our re-
630 sults indicate that the Lesser Antilles subduction zone is uncoupled. Under the paradigm
631 that the degree of interseismic locking correlates with slip during large earthquakes, as
632 shown in an increasing number of studies (e.g., Chlieh et al., 2008; Perfettini et al., 2010;
633 Moreno et al., 2010, Loveless and Meade, 2011), this very low coupling is an indication
634 that very large, Tohoku-like, events are unlikely – or rare.

635 The GPS data also shows a small, but detectable amount of along-arc extension,
636 consistent with geological observations of active normal faulting within the arc. The max-
637 imum extension rate reaches 0.70 mm/yr, which provides an upper bound for long-term
638 slip rates of intra-arc active faults. All $M > 7$ earthquakes in the past ~ 70 years have been
639 normal faulting events, either within the overriding plate or the subducting slab. Although
640 the Lesser Antilles subduction appears to be mechanically uncoupled, implying little to
641 no compressional strain accumulation along the subduction interface, such normal fault-
642 ing events can however be very damaging and are an important hazard source in the Lesser
643 Antilles.

644 The mechanism responsible for the lack of current mechanical coupling at the Lesser
645 Antilles subduction remains elusive, but, as observed in other regions, may be related
646 to the highly hydrated and fractured incoming oceanic lithosphere. As this hydrated oceanic
647 lithosphere subducts, dehydration metamorphic reactions release large amounts of flu-
648 ids that migrate to the plate interface where overpressures are maintained by a low per-
649 meability seal, hence promoting stable creep (Audet and Schwartz, 2013; Moreno et al.,
650 2014). This mechanism is consistent with the high V_p/V_s ratios observed in the central
651 part (Martinique – Antigua) of the Lesser Antilles subduction (Paulatto et al., 2017).

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672 All data used in this work are openly available from the IGS (igsb.jpl.nasa.gov),
 673 UNAVCO (www.unavco.org), and IPGP data center (<http://volobsis.ipgp.fr/>) archives.
 674 Ancillary information necessary to process GPS data, such as precise satellite orbits and
 675 antenna phase center models, is openly available from the IGS (<http://www.igs.org>). Global
 676 SINEX files used here are publicly available at MIT (acc.igs.org/reprocess.html). The
 677 software used to process the GPS data (GAMIT-GLOBK) is openly available at MIT
 678 (<http://geoweb.mit.edu/gg/>). Geodetic coupling models were obtained using a combi-
 679 nation of the CSI (<https://github.com/jolivetr/csi>) and AltTar (<https://github.com/AltTarFramework/altar>)
 680 softwares. Figures were produced using the Generic Mapping Tools software package (Wessel
 681 & Smith, 1998).

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